

# Basic Ground-Water Hydrology

By RALPH C. HEATH

Prepared in cooperation with the  
North Carolina Department of  
Natural Resources and Community  
Development



U.S. GEOLOGICAL SURVEY WATER-SUPPLY PAPER 2220

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DEPARTMENT OF THE INTERIOR  
DONALD PAUL HODEL, Secretary  
  
U.S. GEOLOGICAL SURVEY  
Dallas L. Peck, Director



First printing 1983  
Second printing 1984  
Third printing 1984  
Fourth printing 1987

UNITED STATES GOVERNMENT PRINTING OFFICE: 1987

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For sale by the Books and Open-File Reports Section, U.S. Geological Survey,  
Federal Center, Box 25425, Denver, CO 80225

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**Library of Congress Cataloging in Publication Data**

Heath, Ralph C.  
Basic ground-water hydrology.

(Geological Survey water-supply paper ; 2220)

Bibliography: p. 81

1. Hydrogeology. I. North Carolina. Dept. of Natural  
Resources and Community Development. II. Title.  
III. Series.

GB1003.2.H4 1982 551.49 82-600384

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*Optima*

*Optima  
Bald P's*

## PREFACE

Ground water is one of the Nation's most valuable natural resources. It is the source of about 40 percent of the water used for all purposes exclusive of hydropower generation and electric powerplant cooling.

Surprisingly, for a resource that is so widely used and so important to the health and to the economy of the country, the occurrence of ground water is not only poorly understood but is also, in fact, the subject of many widespread misconceptions. Common misconceptions include the belief that ground water occurs in underground rivers resembling surface streams whose presence can be detected by certain individuals. These misconceptions and others have hampered the development and conservation of ground water and have adversely affected the protection of its quality.

In order for the Nation to receive maximum benefit from its ground-water resource, it is essential that everyone, from the rural homeowner to managers of industrial and municipal water supplies to heads of Federal and State water-regulatory agencies, become more knowledgeable about the occurrence, development, and protection of ground water. This report has been prepared to help meet the needs of these groups, as well as the needs of hydrologists, well drillers, and others engaged in the study and development of ground-water supplies. It consists of 45 sections on the basic elements of ground-water hydrology, arranged in order from the most basic aspects of the subject through a discussion of the methods used to determine the yield of aquifers to a discussion of common problems encountered in the operation of ground-water supplies.

Each section consists of a brief text and one or more drawings or maps that illustrate the main points covered in the text. Because the text is, in effect, an expanded discussion of the illustrations, most of the illustrations are not captioned. However, where more than one drawing is included in a section, each drawing is assigned a number, given in parentheses, and these numbers are inserted at places in the text where the reader should refer to the drawing.

In accordance with U.S. Geological Survey policy to encourage the use of metric units, these units are used in most sections. In the sections dealing with the analysis of aquifer (pumping) test data, equations are given in both consistent units and in the inconsistent inch-pound units still in relatively common use among ground-water hydrologists and well drillers. As an aid to those who are not familiar with metric units and with the conversion of ground-water hydraulic units from inch-pound units to metric units, conversion tables are given on the inside back cover.

Definitions of ground-water terms are given where the terms are first introduced. Because some of these terms will be new to many readers, abbreviated definitions are also given on the inside front cover for convenient reference by those who wish to review the definitions from time to time as they read the text. Finally, for those who need to review some of the simple mathematical operations that are used in ground-water hydrology, a section on numbers, equations, and conversions is included at the end of the text.

**Ralph C. Heath**

# GROUND-WATER HYDROLOGY

*The science of hydrology would be relatively simple if water were unable to penetrate below the earth's surface.*

**Harold E. Thomas**

Ground-water hydrology is the subdivision of the science of hydrology that deals with the occurrence, movement, and quality of water beneath the Earth's surface. It is interdisciplinary in scope in that it involves the application of the physical, biological, and mathematical sciences. It is also a science whose successful application is of critical importance to the welfare of mankind. Because ground-water hydrology deals with the occurrence and movement of water in an almost infinitely complex subsurface environment, it is, in its most advanced state, one of the most complex of the sciences. On the other hand, many of its basic principles and methods can be understood readily by nonhydrologists and used by them in the solution of ground-water problems. The purpose of this report is to present these basic aspects of ground-water hydrology in a form that will encourage more widespread understanding and use.

The ground-water environment is hidden from view except in caves and mines, and the impression that we gain even from these are, to a large extent, misleading. From our observations on the land surface, we form an impression of a "solid" Earth. This impression is not altered very much when we enter a limestone cave and see water flowing in a channel that nature has cut into what appears to be solid rock. In fact, from our observations, both on the land surface and in caves, we are likely to conclude that ground water occurs only in underground rivers and "veins." We do not see the myriad openings that exist between the grains of sand and silt, between particles of clay, or even along the fractures in granite. Consequently, we do not sense the presence of the openings that, in total volume, far exceed the volume of all caves.

R. L. Nace of the U.S. Geological Survey has estimated that the total volume of subsurface openings (which are occupied mainly by water, gas, and petroleum) is on the order of 521,000 km<sup>3</sup> (125,000 mi<sup>3</sup>) beneath the United States alone. If we visualize these openings as forming a continuous cave beneath the entire surface of the United States, its height would be about 57 m (186 ft). The openings, of course, are not equally distributed, the result being that our imaginary cave would range in height from about 3 m (10 ft) beneath the Piedmont Plateau along the eastern seaboard to about 2,500 m (8,200 ft) beneath the Mississippi Delta. The important point to be gained from this discussion is that the total volume of openings beneath the surface of the United States, and other land areas of the world, is very large.

Most subsurface openings contain water, and the importance of this water to mankind can be readily demonstrated by comparing its volume with the volumes of water in other parts of the hydrosphere.<sup>1</sup> Estimates of the volumes of water in the hydrosphere have been made by the Russian hydrologist M. I. L'vovich and are given in a book recently translated into English. Most water, including that in the oceans and in

the deeper subsurface openings, contains relatively large concentrations of dissolved minerals and is not readily usable for essential human needs. We will, therefore, concentrate in this discussion only on freshwater. The accompanying table contains L'vovich's estimates of the freshwater in the hydrosphere. Not surprisingly, the largest volume of freshwater occurs as ice in glaciers. On the other hand, many people impressed by the "solid" Earth are surprised to learn that about 14 percent of all freshwater is ground water and that, if only water is considered, 94 percent is ground water.

Ground-water hydrology, as noted earlier, deals not only with the occurrence of underground water but also with its movement. Contrary to our impressions of rapid movement as we observe the flow of streams in caves, the movement of most ground water is exceedingly slow. The truth of this observation becomes readily apparent from the table, which shows, in the last column, the rate of water exchange or the time required to replace the water now contained in the listed parts of the hydrosphere. It is especially important to note that the rate of exchange of 280 years for fresh ground water is about 1/9,000 the rate of exchange of water in rivers.

Subsurface openings large enough to yield water in a usable quantity to wells and springs underlie nearly every place on the land surface and thus make ground water one of the most widely available natural resources. When this fact and the fact that ground water also represents the largest reservoir of freshwater readily available to man are considered together, it is obvious that the value of ground water, in terms of both economics and human welfare, is incalculable. Consequently, its sound development, diligent conservation, and consistent protection from pollution are important concerns of everyone. These concerns can be translated into effective action only by increasing our knowledge of the basic aspects of ground-water hydrology.

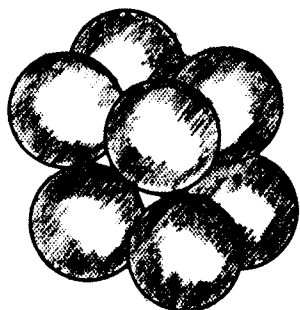
## FRESHWATER OF THE HYDROSPHERE AND ITS RATE OF EXCHANGE

[Modified from L'vovich (1979), tables 2 and 10]

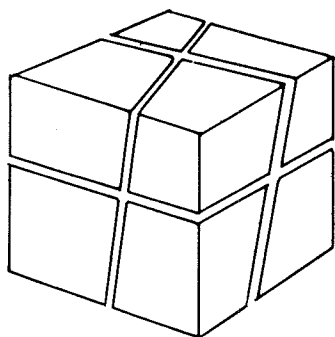
Parts of the hydrosphere	Volume of freshwater		Share in total volume of freshwater (percent)	Rate of water exchange (yr)
	km <sup>3</sup>	mi <sup>3</sup>		
Ice sheets and glaciers -----	24,000,000	5,800,000	84.945	8,000
Ground water --	4,000,000	960,000	14.158	280
Lakes and reservoirs ----	155,000	37,000	.549	7
Soil moisture ---	83,000	20,000	.294	1
Vapors in the atmosphere --	14,000	3,400	.049	.027
River water ----	1,200	300	.004	.031
Total -----	28,253,200	6,820,700	100.000	

<sup>1</sup>The hydrosphere is the term used to refer to the waters of the Earth and, in its broadest usage, includes all water, water vapor, and ice regardless of whether they occur beneath, on, or above the Earth's surface.

# ROCKS AND WATER



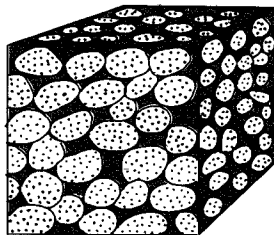
POROUS MATERIAL



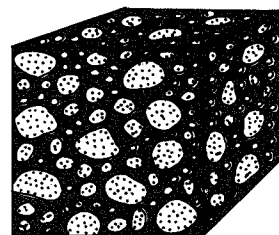
FRACTURED ROCK

( 1 )

## PRIMARY OPENINGS

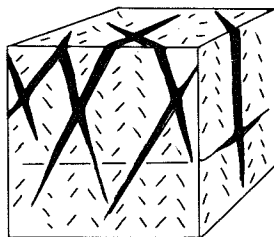


WELL-SORTED SAND

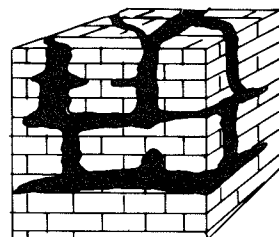


POORLY-SORTED SAND

## SECONDARY OPENINGS



FRACTURES IN  
GRANITE



CAVERNS IN  
LIMESTONE

( 2 )

Most of the rocks near the Earth's surface are composed of both solids and voids, as sketch 1 shows. The solid part is, of course, much more obvious than the voids, but, without the voids, there would be no water to supply wells and springs.

Water-bearing rocks consist either of unconsolidated (soil-like) deposits or consolidated rocks. The Earth's surface in most places is formed by soil and by unconsolidated deposits that range in thickness from a few centimeters near outcrops of consolidated rocks to more than 12,000 m beneath the delta of the Mississippi River. The unconsolidated deposits are underlain everywhere by consolidated rocks.

Most *unconsolidated deposits* consist of material derived from the disintegration of consolidated rocks. The material consists, in different types of unconsolidated deposits, of particles of rocks or minerals ranging in size from fractions of a millimeter (clay size) to several meters (boulders). Unconsolidated deposits important in ground-water hydrology include,

in order of increasing grain size, clay, silt, sand, and gravel. An important group of unconsolidated deposits also includes fragments of shells of marine organisms.

*Consolidated rocks* consist of mineral particles of different sizes and shapes that have been welded by heat and pressure or by chemical reactions into a solid mass. Such rocks are commonly referred to in ground-water reports as *bedrock*. They include sedimentary rocks that were originally unconsolidated and igneous rocks formed from a molten state. Consolidated sedimentary rocks important in ground-water hydrology include limestone, dolomite, shale, siltstone, sandstone, and conglomerate. Igneous rocks include granite and basalt.

There are different kinds of voids in rocks, and it is sometimes useful to be aware of them. If the voids were formed at the same time as the rock, they are referred to as *primary openings* (2). The pores in sand and gravel and in other unconsolidated deposits are primary openings. The lava tubes and other openings in basalt are also primary openings.

If the voids were formed after the rock was formed, they are referred to as *secondary openings* (2). The fractures in granite and in consolidated sedimentary rocks are secondary openings. Voids in limestone, which are formed as ground water slowly dissolves the rock, are an especially important type of secondary opening.

It is useful to introduce the topic of rocks and water by dealing with unconsolidated deposits on one hand and with

consolidated rocks on the other. It is important to note, however, that many sedimentary rocks that serve as sources of ground water fall between these extremes in a group of *semi-consolidated rocks*. These are rocks in which openings include both pores and fractures—in other words, both primary and secondary openings. Many limestones and sandstones that are important sources of ground water are semiconsolidated.

# UNDERGROUND WATER

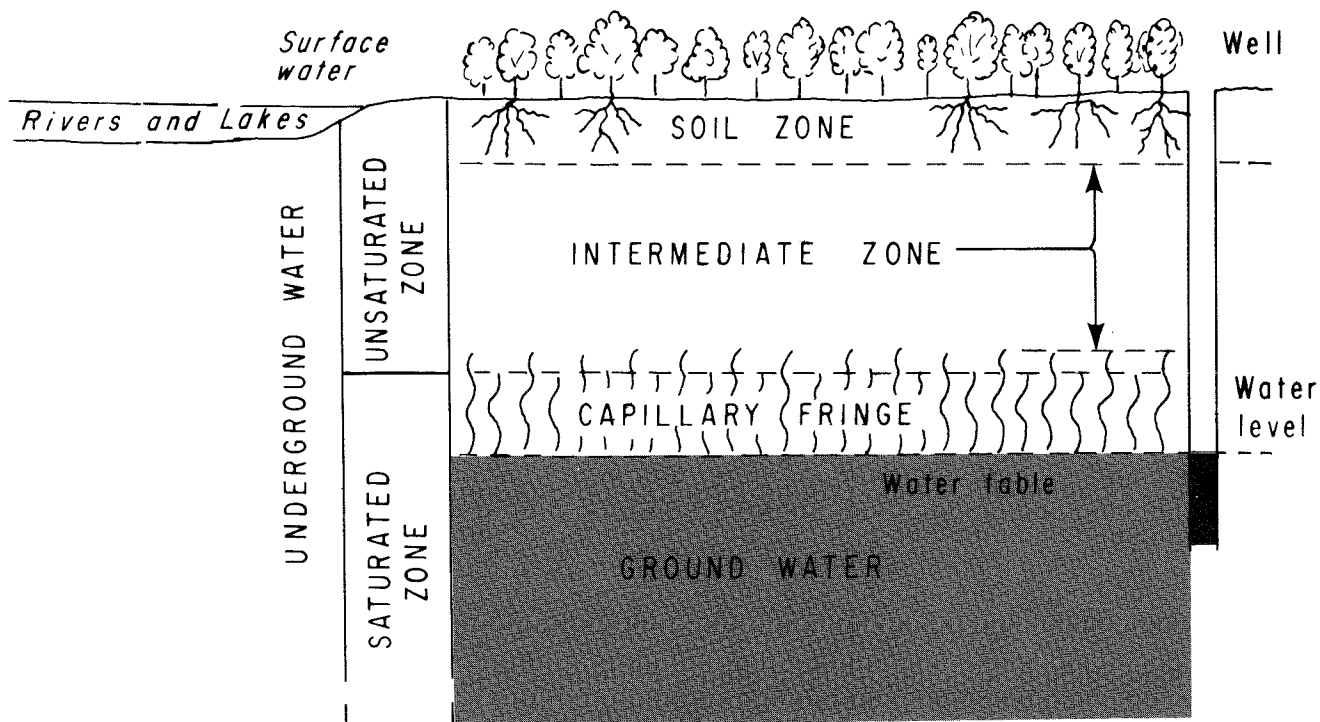
All water beneath the land surface is referred to as *underground water* (or subsurface water). The equivalent term for water on the land surface is *surface water*. Underground water occurs in two different zones. One zone, which occurs immediately below the land surface in most areas, contains both water and air and is referred to as the *unsaturated zone*. The unsaturated zone is almost invariably underlain by a zone in which all interconnected openings are full of water. This zone is referred to as the *saturated zone*.

Water in the saturated zone is the only underground water that is available to supply wells and springs and is the only water to which the name *ground water* is correctly applied. Recharge of the saturated zone occurs by percolation of water from the land surface through the unsaturated zone. The unsaturated zone is, therefore, of great importance to ground-water hydrology. This zone may be divided usefully into three parts: the soil zone, the intermediate zone, and the upper part of the capillary fringe.

The soil zone extends from the land surface to a maximum depth of a meter or two and is the zone that supports plant growth. It is crisscrossed by living roots, by voids left by

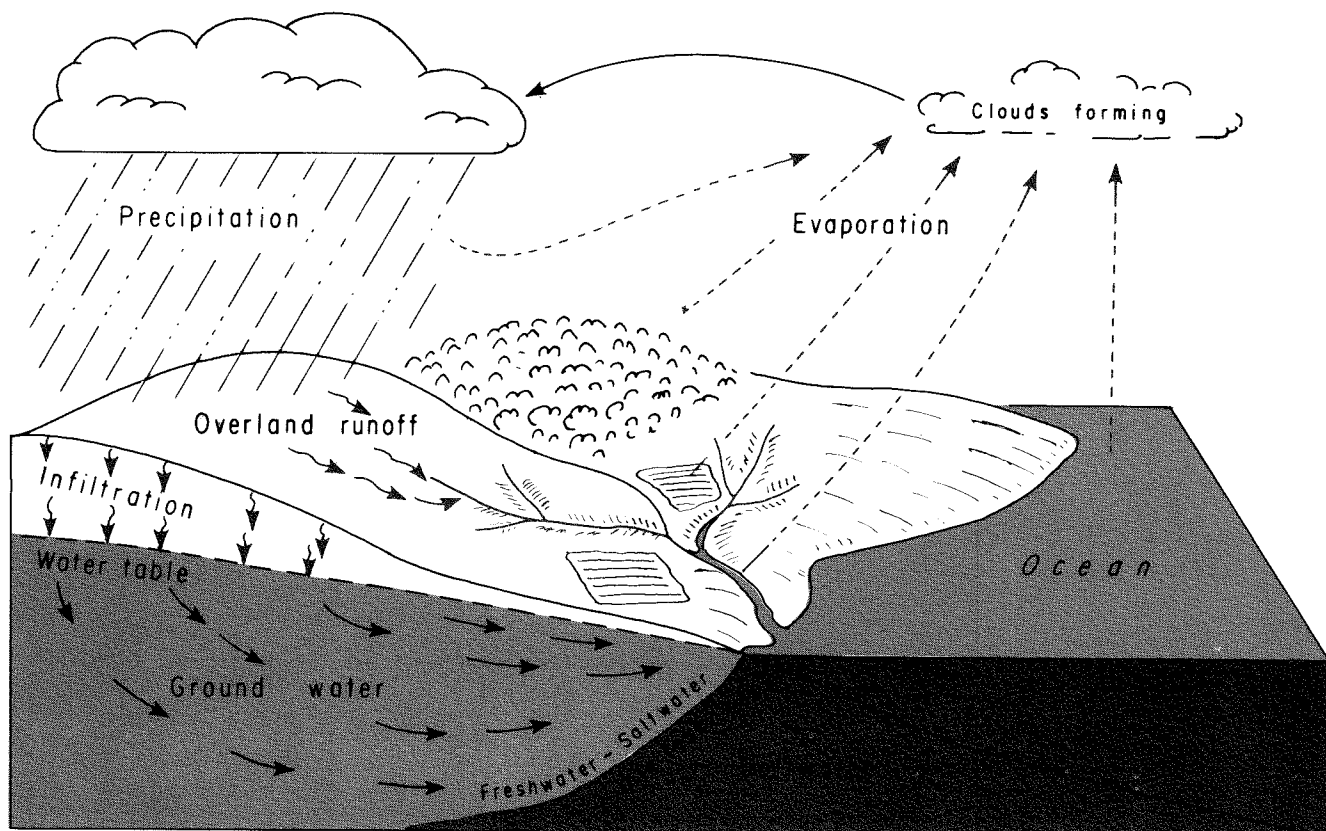
decayed roots of earlier vegetation, and by animal and worm burrows. The porosity and permeability of this zone tend to be higher than those of the underlying material. The soil zone is underlain by the *intermediate zone*, which differs in thickness from place to place depending on the thickness of the soil zone and the depth to the capillary fringe.

The lowest part of the unsaturated zone is occupied by the *capillary fringe*, the subzone between the unsaturated and saturated zones. The capillary fringe results from the attraction between water and rocks. As a result of this attraction, water clings as a film on the surface of rock particles and rises in small-diameter pores against the pull of gravity. Water in the capillary fringe and in the overlying part of the unsaturated zone is under a negative hydraulic pressure—that is, it is under a pressure less than the atmospheric (barometric) pressure. The *water table* is the level in the saturated zone at which the hydraulic pressure is equal to atmospheric pressure and is represented by the water level in unused wells. Below the water table, the hydraulic pressure increases with increasing depth.





# HYDROLOGIC CYCLE



The term *hydrologic cycle* refers to the constant movement of water above, on, and below the Earth's surface. The concept of the hydrologic cycle is central to an understanding of the occurrence of water and the development and management of water supplies.

Although the hydrologic cycle has neither a beginning nor an end, it is convenient to discuss its principal features by starting with evaporation from vegetation, from exposed moist surfaces including the land surface, and from the ocean. This moisture forms clouds, which return the water to the land surface or oceans in the form of precipitation.

Precipitation occurs in several forms, including rain, snow, and hail, but only rain is considered in this discussion. The first rain wets vegetation and other surfaces and then begins to infiltrate into the ground. *Infiltration* rates vary widely, depending on land use, the character and moisture content of the soil, and the intensity and duration of precipitation, from possibly as much as 25 mm/hr in mature forests on sandy soils to a few millimeters per hour in clayey and silty soils to zero in paved areas. When and if the rate of precipitation exceeds the rate of infiltration, *overland flow* occurs.

The first infiltration replaces soil moisture, and, thereafter, the excess percolates slowly across the intermediate zone to the zone of saturation. Water in the zone of saturation moves

downward and laterally to sites of ground-water discharge such as springs on hillsides or seeps in the bottoms of streams and lakes or beneath the ocean.

Water reaching streams, both by overland flow and from ground-water discharge, moves to the sea, where it is again evaporated to perpetuate the cycle.

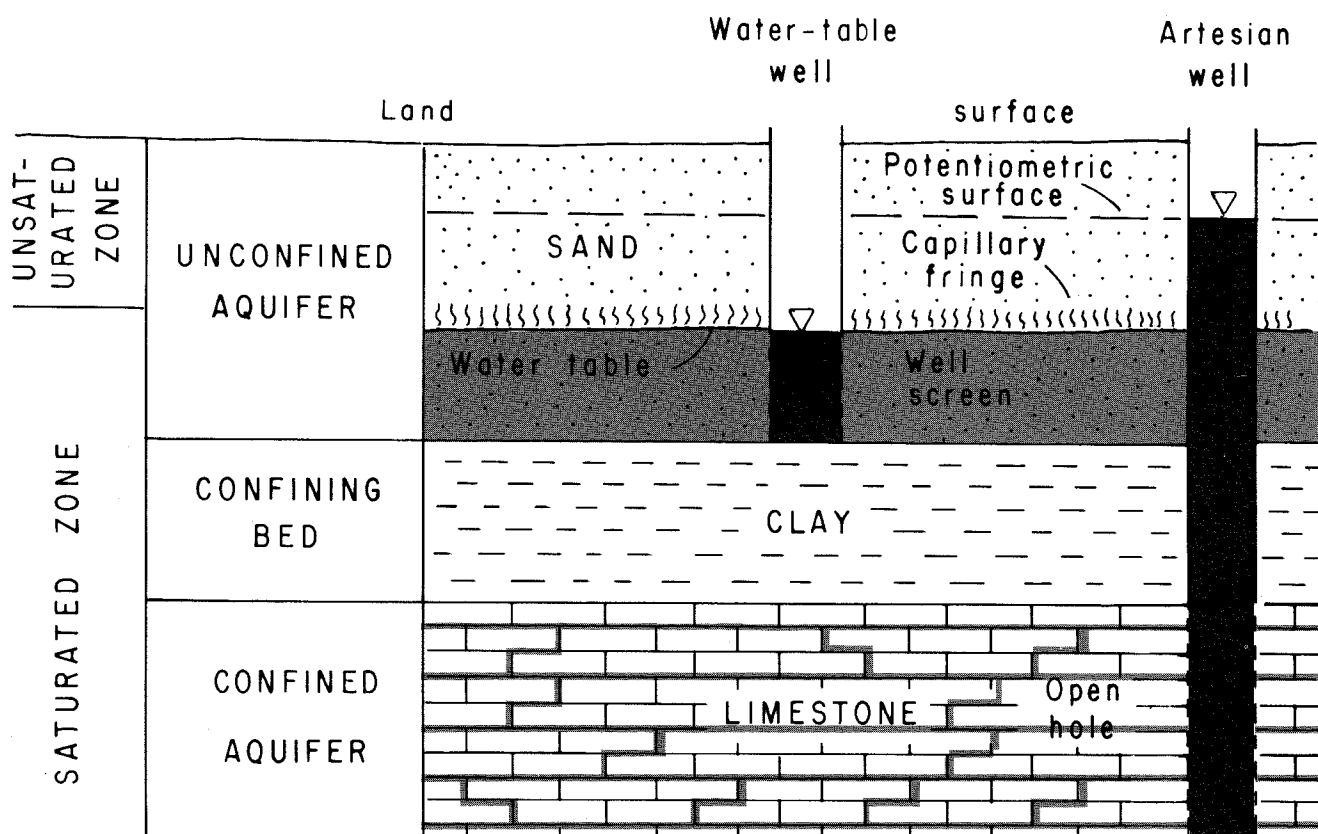
Movement is, of course, the key element in the concept of the hydrologic cycle. Some "typical" rates of movement are shown in the following table, along with the distribution of the Earth's water supply.

## RATE OF MOVEMENT AND DISTRIBUTION OF WATER

[Adapted from L'vovich (1979), table 1]

Location	Rate of movement	Distribution of Earth's water supply (percent)
Atmosphere ---	100's of kilometers per day	0.001
Water on land surface -----	10's of kilometers per day	.019
Water below the land surface --	Meters per year	4.12
Ice caps and glaciers -----	Meters per day	1.65
Oceans -----	--	93.96

# AQUIFERS AND CONFINING BEDS



From the standpoint of ground-water occurrence, all rocks that underlie the Earth's surface can be classified either as aquifers or as confining beds. An *aquifer* is a rock unit that will yield water in a usable quantity to a well or spring. (In geologic usage, "rock" includes unconsolidated sediments.) A *confining bed* is a rock unit having very low hydraulic conductivity that restricts the movement of ground water either into or out of adjacent aquifers.

Ground water occurs in aquifers under two different conditions. Where water only partly fills an aquifer, the upper surface of the saturated zone is free to rise and decline. The water in such aquifers is said to be *unconfined*, and the aquifers are referred to as *unconfined aquifers*. Unconfined aquifers are also widely referred to as *water-table aquifers*.

Where water completely fills an aquifer that is overlain by a confining bed, the water in the aquifer is said to be *confined*. Such aquifers are referred to as *confined aquifers* or as *artesian aquifers*.

Wells open to unconfined aquifers are referred to as *water-table wells*. The water level in these wells indicates the position of the water table in the surrounding aquifer.

Wells drilled into confined aquifers are referred to as *artesian wells*. The water level in artesian wells stands at some height above the top of the aquifer but not necessarily above the land surface. If the water level in an artesian well stands above the land surface, the well is a *flowing artesian well*. The water level in tightly cased wells open to a confined aquifer stands at the level of the *potentiometric surface* of the aquifer.

# POROSITY

The ratio of openings (voids) to the total volume of a soil or rock is referred to as its *porosity*. Porosity is expressed either as a decimal fraction or as a percentage. Thus,

$$n = \frac{V_t - V_s}{V_t} = \frac{V_v}{V_t}$$

where *n* is porosity as a decimal fraction, *V<sub>t</sub>* is the total volume of a soil or rock sample, *V<sub>s</sub>* is the volume of solids in the sample, and *V<sub>v</sub>* is the volume of openings (voids).

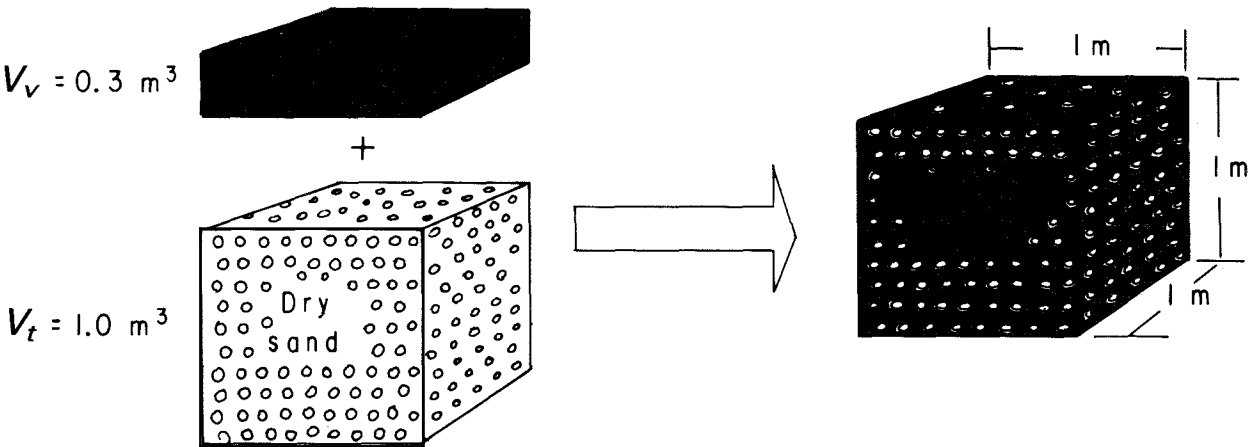
If we multiply the porosity determined with the equation by 100, the result is porosity expressed as a percentage.

Soils are among the most porous of natural materials because soil particles tend to form loose clumps and because of the presence of root holes and animal burrows. Porosity of unconsolidated deposits depends on the range in grain size (sorting) and on the shape of the rock particles but not on their size. Fine-grained materials tend to be better sorted and, thus, tend to have the largest porosities.

## SELECTED VALUES OF POROSITY

[Values in percent by volume]

Material	Primary openings	Secondary openings
Equal-size spheres (marbles):		
Loosest packing -----	48	--
Tightest packing -----	26	--
Soil -----	55	--
Clay -----	50	--
Sand -----	25	--
Gravel -----	20	--
Limestone -----	10	10
Sandstone (semiconsolidated) ----	10	1
Granite -----	--	.1
Basalt (young) -----	10	1



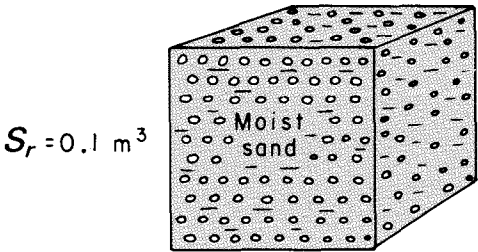
$$\text{Porosity } (n) = \frac{\text{Volume of voids } (V_v)}{\text{Total volume } (V_t)} = \frac{0.3 \text{ m}^3}{1.0 \text{ m}^3} = 0.30$$

# SPECIFIC YIELD AND SPECIFIC RETENTION

Porosity is important in ground-water hydrology because it tells us the maximum amount of water that a rock can contain when it is saturated. However, it is equally important to know that only a part of this water is available to supply a well or a spring.

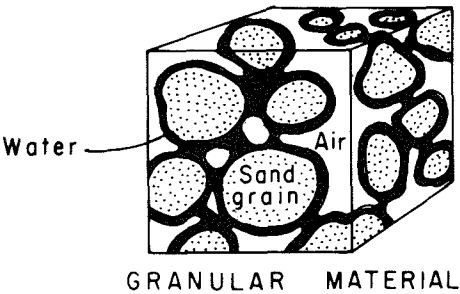
Hydrologists divide water in storage in the ground into the

part that will drain under the influence of gravity (called *specific yield*) (1) and the part that is retained as a film on rock surfaces and in very small openings (called *specific retention*) (2). The physical forces that control specific retention are the same forces involved in the thickness and moisture content of the capillary fringe.

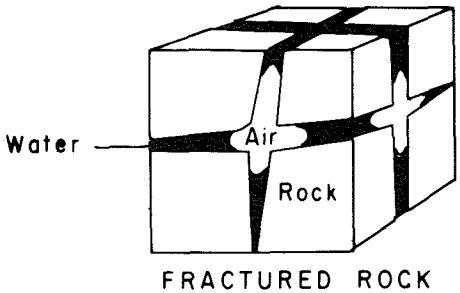


$$n = S_y + S_r = \frac{0.2 \text{ m}^3}{1 \text{ m}^3} + \frac{0.1 \text{ m}^3}{1 \text{ m}^3} = 0.30$$

(1)



Water retained as a film on rock surfaces and in capillary-size openings after gravity drainage.



(2)

Specific yield tells how much water is available for man's use, and specific retention tells how much water remains in the rock after it is drained by gravity. Thus,

$$n = S_y + S_r$$

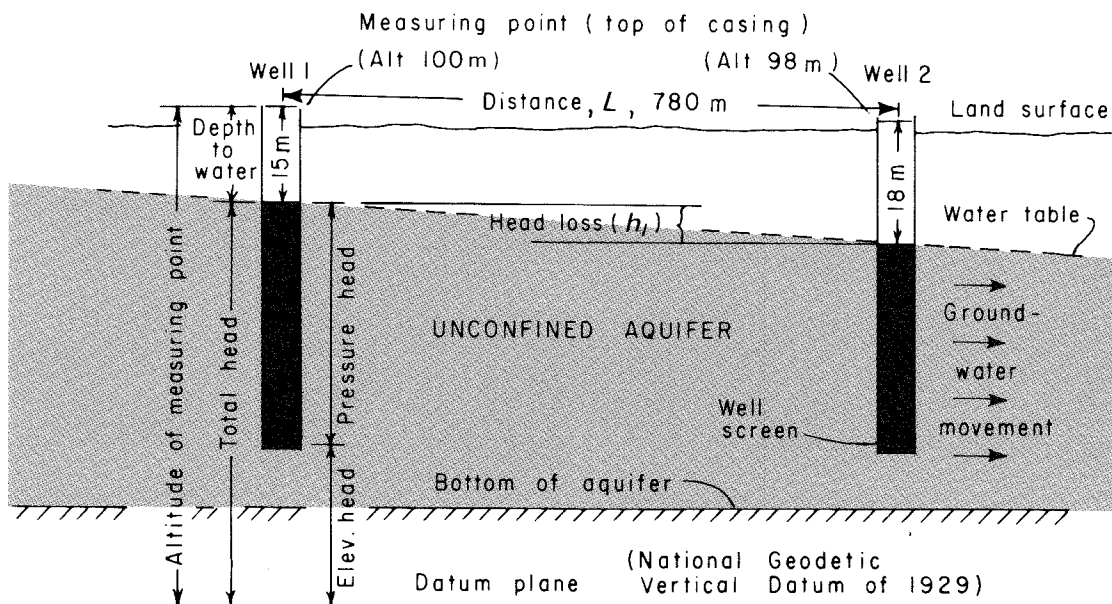
$$S_y = \frac{V_d}{V_t} \qquad S_r = \frac{V_r}{V_t}$$

where  $n$  is porosity,  $S_y$  is specific yield,  $S_r$  is specific retention,  $V_d$  is the volume of water than drains from a total volume of  $V_t$ ,  $V_r$  is the volume of water retained in a total volume of  $V_t$ , and  $V_t$  is total volume of a soil or rock sample.

SELECTED VALUES OF POROSITY, SPECIFIC YIELD,  
AND SPECIFIC RETENTION  
[Values in percent by volume]

Material	Porosity	Specific yield	Specific retention
Soil -----	55	40	15
Clay -----	50	2	48
Sand -----	25	22	3
Gravel -----	20	19	1
Limestone -----	20	18	2
Sandstone (semiconsolidated)	11	6	5
Granite -----	.1	.09	.01
Basalt (young) -----	11	8	3

# HEADS AND GRADIENTS



(1)

The depth to the water table has an important effect on use of the land surface and on the development of water supplies from unconfined aquifers (1). Where the water table is at a shallow depth, the land may become "waterlogged" during wet weather and unsuitable for residential and many other uses. Where the water table is at great depth, the cost of constructing wells and pumping water for domestic needs may be prohibitively expensive.

The direction of the slope of the water table is also important because it indicates the direction of ground-water movement (1). The position and the slope of the water table (or of the potentiometric surface of a confined aquifer) is determined by measuring the position of the water level in wells from a fixed point (a measuring point) (1). (See "Measurements of Water levels and Pumping Rates.") To utilize these measurements to determine the slope of the water table, the position of the water table at each well must be determined relative to a *datum plane* that is common to all the wells. The datum plane most widely used is the National Geodetic Vertical Datum of 1929 (also commonly referred to as "sea level") (1).

If the depth to water in a nonflowing well is subtracted from the altitude of the measuring point, the result is the *total head* at the well. Total head, as defined in fluid mechanics, is composed of *elevation head*, *pressure head*, and *velocity head*. Because ground water moves relatively slowly, velocity head can be ignored. Therefore, the total head at an observation well involves only two components: elevation head and pressure head (1). Ground water moves in the direction of decreasing total head, which may or may not be in the direction of decreasing pressure head.

The equation for total head ( $h_t$ ) is

$$h_t = z + h_p$$

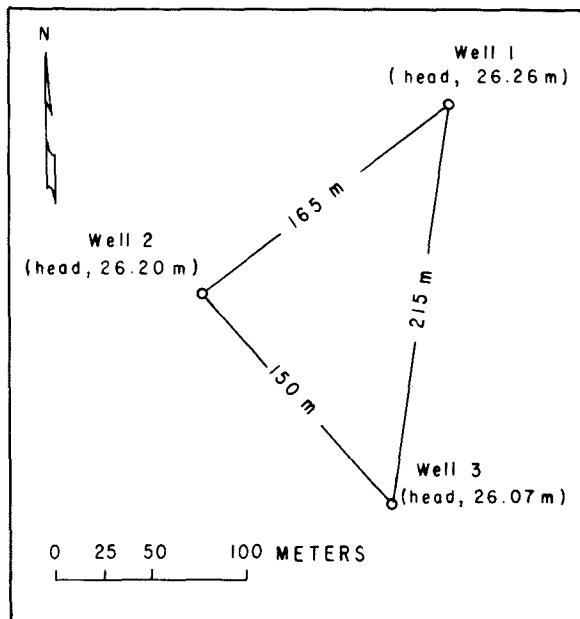
where  $z$  is elevation head and is the distance from the datum plane to the point where the pressure head  $h_p$  is determined.

All other factors being constant, the rate of ground-water movement depends on the *hydraulic gradient*. The hydraulic gradient is the change in head per unit of distance in a given direction. If the direction is not specified, it is understood to be in the direction in which the maximum rate of decrease in head occurs.

If the movement of ground water is assumed to be in the plane of sketch 1—in other words, if it moves from well 1 to well 2—the hydraulic gradient can be calculated from the information given on the drawing. The hydraulic gradient is  $h_L/L$ , where  $h_L$  is the head loss between wells 1 and 2 and  $L$  is the horizontal distance between them, or

$$\frac{h_L}{L} = \frac{(100 \text{ m} - 15 \text{ m}) - (98 \text{ m} - 18 \text{ m})}{780 \text{ m}} = \frac{85 \text{ m} - 80 \text{ m}}{780 \text{ m}} = \frac{5 \text{ m}}{780 \text{ m}}$$

When the hydraulic gradient is expressed in consistent units, as it is in the above example in which both the numerator and the denominator are in meters, any other consistent units of length can be substituted without changing the value of the gradient. Thus, a gradient of 5 ft/780 ft is the same as a gradient of 5 m/780 m. It is also relatively common to express hydraulic gradients in inconsistent units such as meters per



( 2 )

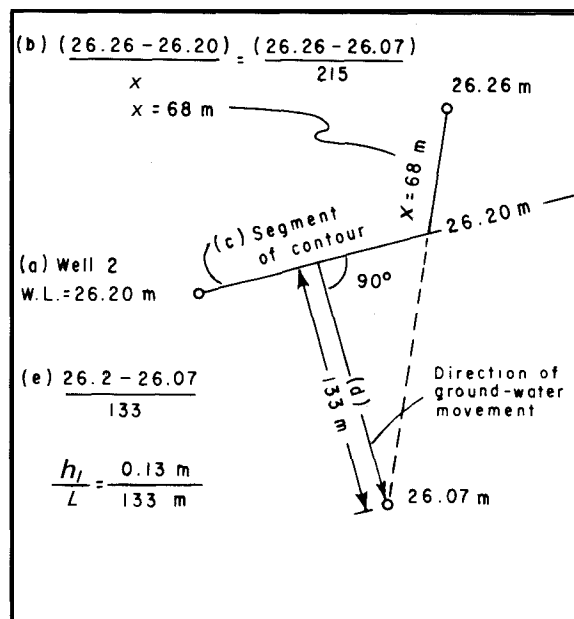
kilometer or feet per mile. A gradient of 5 m/780 m can be converted to meters per kilometer as follows:

$$\left( \frac{5 \text{ m}}{780 \text{ m}} \right) \times \left( \frac{1,000 \text{ m}}{\text{km}} \right) = 6.4 \text{ m km}^{-1}$$

Both the direction of ground-water movement and the hydraulic gradient can be determined if the following data are available for three wells located in any triangular arrangement such as that shown on sketch 2:

1. The relative geographic position of the wells.
2. The distance between the wells.
3. The total head at each well.

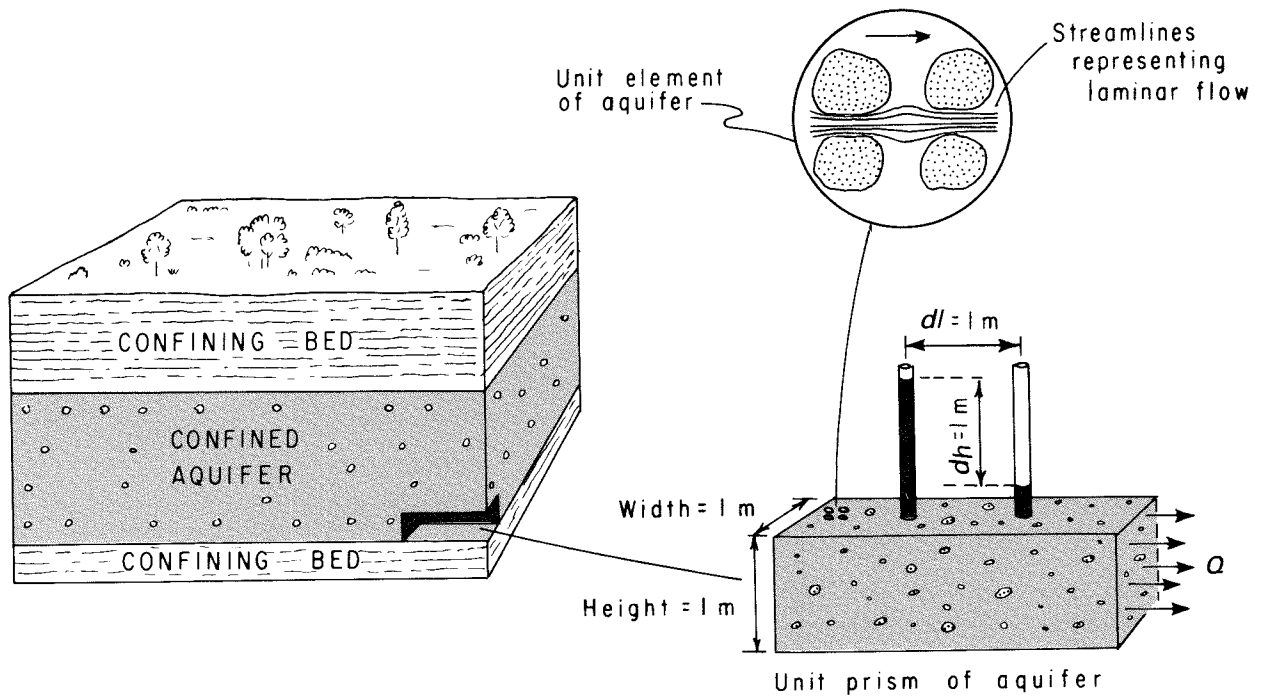
Steps in the solution are outlined below and illustrated in sketch 3:



( 3 )

- a. Identify the well that has the intermediate water level (that is, neither the highest head nor the lowest head).
- b. Calculate the position between the well having the highest head and the well having the lowest head at which the head is the same as that in the intermediate well.
- c. Draw a straight line between the intermediate well and the point identified in step b as being between the well having the highest head and that having the lowest head. This line represents a segment of the water-level contour along which the total head is the same as that in the intermediate well.
- d. Draw a line perpendicular to the water-level contour and through either the well with the highest head or the well with the lowest head. This line parallels the direction of ground-water movement.
- e. Divide the difference between the head of the well and that of the contour by the distance between the well and the contour. The answer is the hydraulic gradient.

# HYDRAULIC CONDUCTIVITY



(1)

Aquifers transmit water from recharge areas to discharge areas and thus function as porous conduits (or pipelines filled with sand or other water-bearing material). The factors controlling ground-water movement were first expressed in the form of an equation by Henry Darcy, a French engineer, in 1856. Darcy's law is

$$Q = KA \left( \frac{dh}{dl} \right) \quad (1)$$

where  $Q$  is the quantity of water per unit of time;  $K$  is the hydraulic conductivity and depends on the size and arrangement of the water-transmitting openings (pores and fractures) and on the dynamic characteristics of the fluid (water) such as kinematic viscosity, density, and the strength of the gravitational field;  $A$  is the cross-sectional area, at a right angle to the flow direction, through which the flow occurs; and  $dh/dl$  is the hydraulic gradient.<sup>1</sup>

Because the quantity of water ( $Q$ ) is directly proportional to the hydraulic gradient ( $dh/dl$ ), we say that ground-water flow is *laminar*—that is, water particles tend to follow discrete streamlines and not to mix with particles in adjacent streamlines (1). (See "Ground-Water Flow Nets.")

<sup>1</sup>Where hydraulic gradient is discussed as an independent entity, as it is in "Heads and Gradients," it is shown symbolically as  $h_l/L$  and is referred to as head loss per unit of distance. Where hydraulic gradient appears as one of the factors in an equation, as it does in equation 1, it is shown symbolically as  $dh/dl$  to be consistent with other ground-water literature. The gradient  $dh/dl$  indicates that the unit distance is reduced to as small a value as one can imagine, in accordance with the concepts of differential calculus.

If we rearrange equation 1 to solve for  $K$ , we obtain

$$K = \frac{Qdl}{Adh} = \frac{(m^3 d^{-1})(m)}{(m^2)(m)} = \frac{m}{d} \quad (2)$$

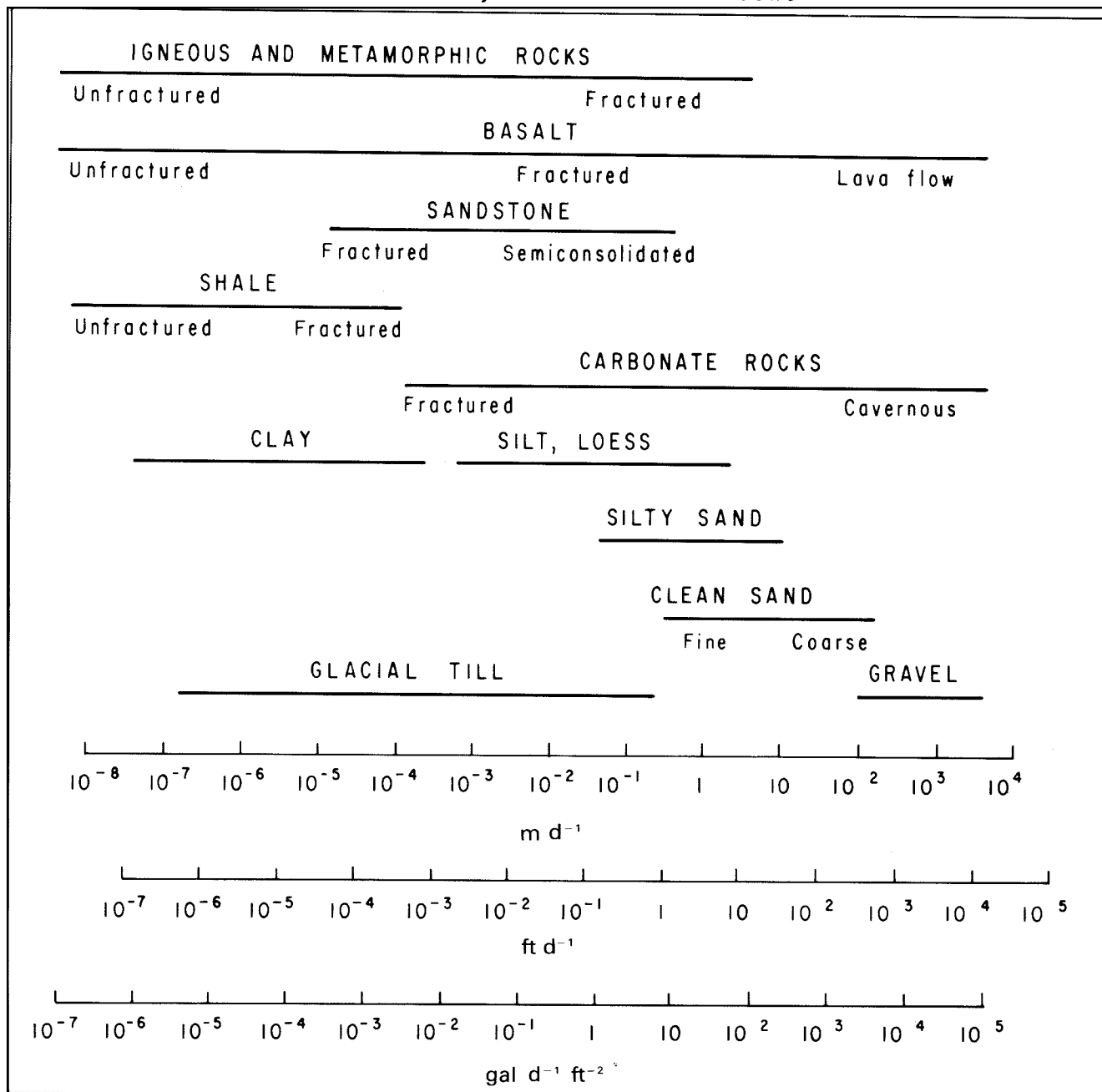
Thus, the units of hydraulic conductivity are those of velocity (or distance divided by time). It is important to note from equation 2, however, that the factors involved in the definition of hydraulic conductivity include the volume of water ( $Q$ ) that will move in a unit of time (commonly, a day) under a unit hydraulic gradient (such as a meter per meter) through a unit area (such as a square meter). These factors are illustrated in sketch 1. Expressing hydraulic conductivity in terms of a unit gradient, rather than of an actual gradient at some place in an aquifer, permits ready comparison of values of hydraulic conductivity for different rocks.

Hydraulic conductivity replaces the term "field coefficient of permeability" and should be used in referring to the water-transmitting characteristic of material in quantitative terms. It is still common practice to refer in qualitative terms to "permeable" and "impermeable" material.

The hydraulic conductivity of rocks ranges through 12 orders of magnitude (2). There are few physical parameters whose values range so widely. Hydraulic conductivity is not only different in different types of rocks but may also be different from place to place in the same rock. If the hydraulic conductivity is essentially the same in any area, the aquifer in



# Hydraulic Conductivity of Selected Rocks



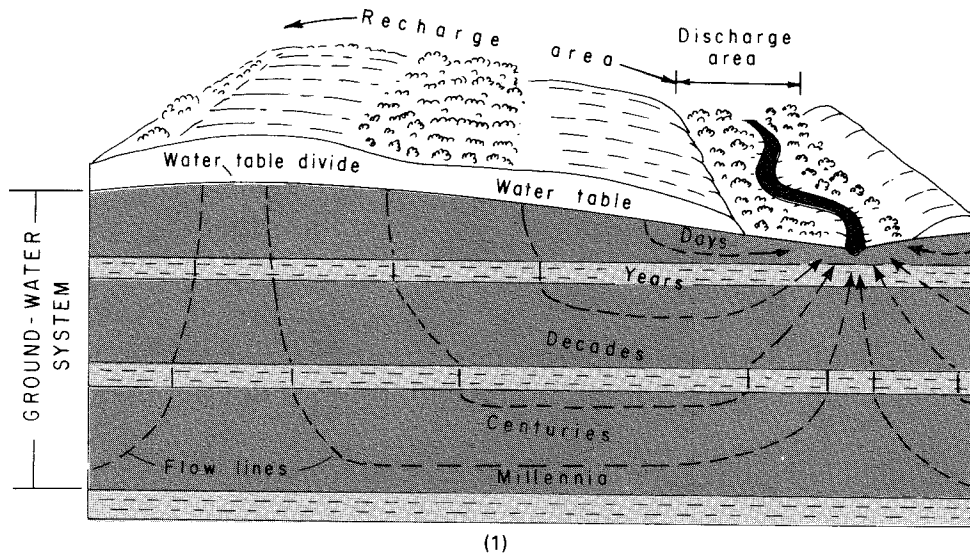
(2)

that area is said to be *homogeneous*. If, on the other hand, the hydraulic conductivity differs from one part of the area to another, the aquifer is said to be *heterogeneous*.

Hydraulic conductivity may also be different in different directions at any place in an aquifer. If the hydraulic conductivity is essentially the same in all directions, the aquifer is said to be *isotropic*. If it is different in different directions, the aquifer is said to be *anisotropic*.

Although it is convenient in many mathematical analyses of ground-water flow to assume that aquifers are both homogeneous and isotropic, such aquifers are rare, if they exist at all. The condition most commonly encountered is for hydraulic conductivity in most rocks and especially in unconsolidated deposits and in flat-lying consolidated sedimentary rocks to be larger in the horizontal direction than it is in the vertical direction.

# FUNCTIONS OF GROUND-WATER SYSTEMS



The aquifers and confining beds that underlie any area comprise the *ground-water system* of the area (1). Hydraulically, this system serves two functions: it stores water to the extent of its porosity, and it transmits water from recharge areas to discharge areas. Thus, a ground-water system serves as both a reservoir and a conduit. With the exception of cavernous limestones, lava flows, and coarse gravels, ground-water systems are more effective as reservoirs than as conduits.

Water enters ground-water systems in *recharge areas* and moves through them, as dictated by hydraulic gradients and hydraulic conductivities, to *discharge areas* (1).

The identification of recharge areas is becoming increasingly important because of the expanding use of the land surface for waste disposal. In the humid part of the country, recharge occurs in all interstream areas—that is, in all areas except along streams and their adjoining flood plains (1). The streams and flood plains are, under most conditions, discharge areas.

In the drier part (western half) of the conterminous United States, recharge conditions are more complex. Most recharge occurs in the mountain ranges, on alluvial fans that border the mountain ranges, and along the channels of major streams where they are underlain by thick and permeable alluvial deposits.

Recharge rates are generally expressed in terms of volume (such as cubic meters or gallons) per unit of time (such as a day or a year) per unit of area (such as a square kilometer, a square mile, or an acre). When these units are reduced to their simplest forms, the result is recharge expressed as a depth of water on the land surface per unit of time. Recharge varies from year to year, depending on the amount of precipitation, its seasonal distribution, air temperature, land use, and other factors. Relative to land use, recharge rates in forests are much higher than those in cities.

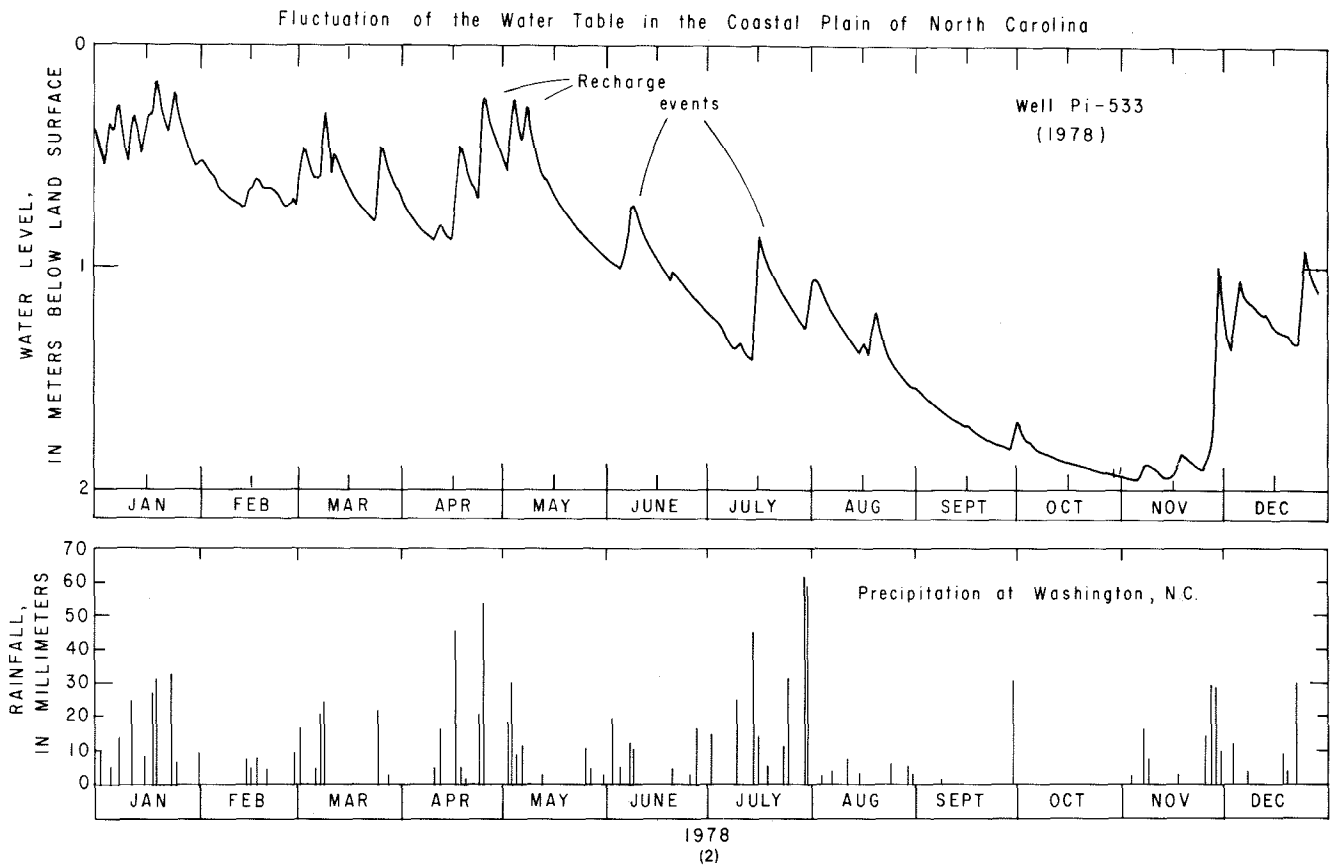
Annual recharge rates range, in different parts of the coun-

try, from essentially zero in desert areas to about  $600 \text{ mm yr}^{-1}$  ( $1,600 \text{ m}^3 \text{ km}^{-2} \text{ d}^{-1}$  or  $1.1 \times 10^6 \text{ gal mi}^{-2} \text{ d}^{-1}$ ) in the rural areas on Long Island and in other rural areas in the East that are underlain by very permeable soils.

The rate of movement of ground water from recharge areas to discharge areas depends on the hydraulic conductivities of the aquifers and confining beds, if water moves downward into other aquifers, and on the hydraulic gradients. (See "Ground-Water Velocity.") A convenient way of showing the rate is in terms of the time required for ground water to move from different parts of a recharge area to the nearest discharge area. The time ranges from a few days in the zone adjacent to the discharge area to thousands of years (millennia) for water that moves from the central part of some recharge areas through the deeper parts of the ground-water system (1).

Natural discharge from ground-water systems includes not only the flow of springs and the seepage of water into stream channels or wetlands but also evaporation from the upper part of the capillary fringe, where it occurs within a meter or so of the land surface. Large amounts of water are also withdrawn from the capillary fringe and the zone of saturation by plants during the growing season. Thus, discharge areas include not only the channels of perennial streams but also the adjoining flood plains and other low-lying areas.

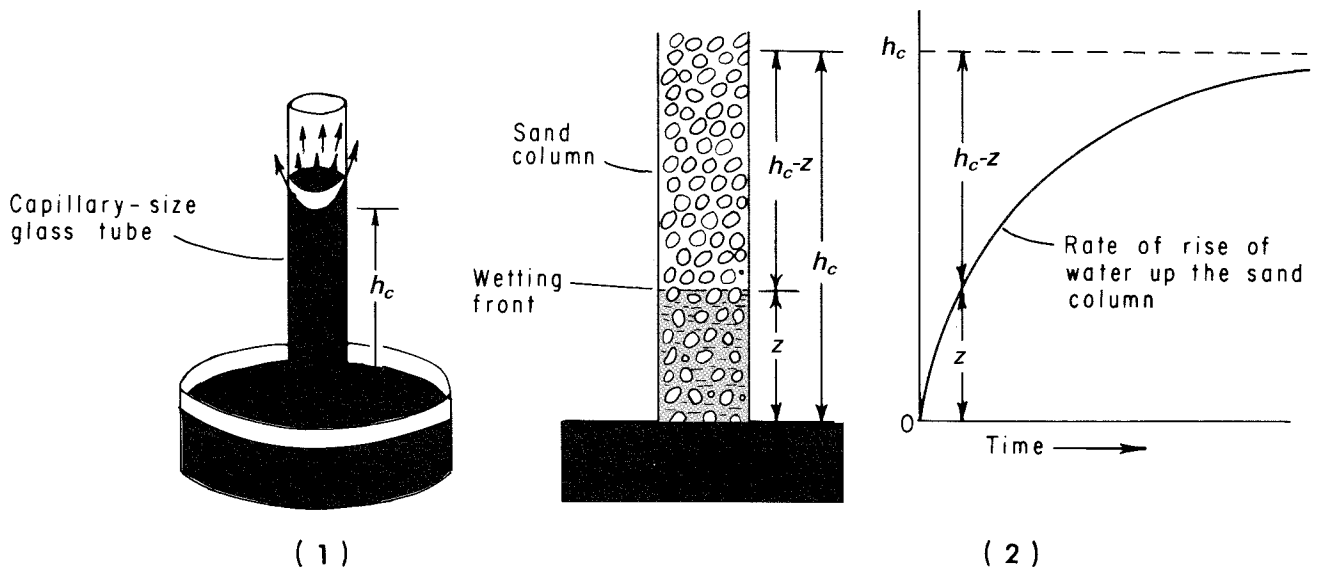
One of the most significant differences between recharge areas and discharge areas is that the areal extent of discharge areas is invariably much smaller than that of recharge areas. This size difference shows, as we would expect, that discharge areas are more "efficient" than recharge areas. Recharge involves unsaturated movement of water in the vertical direction; in other words, movement is in the direction in which the hydraulic conductivity is generally the lowest. Discharge, on the other hand, involves saturated movement, much of it in the horizontal direction—that is, in the direction of the largest hydraulic conductivity.



Another important aspect of recharge and discharge involves timing. Recharge occurs during and immediately following periods of precipitation and thus is intermittent (2). Discharge, on the other hand, is a continuous process as long as ground-water heads are above the level at which discharge occurs. However, between periods of recharge, ground-water heads decline, and the rate of discharge also declines. Most recharge of ground-water systems occurs during late fall,

winter, and early spring, when plants are dormant and evaporation rates are small. These aspects of recharge and discharge are apparent from graphs showing the fluctuation of the water level in observation wells, such as the one shown in sketch 2. The occasional lack of correlation, especially in the summer, between the precipitation and the rise in water level is due partly to the distance of 20 km between the weather station and the well.

# CAPILLARITY AND UNSATURATED FLOW



Most recharge of ground-water systems occurs during the percolation of water across the unsaturated zone. The movement of water in the unsaturated zone is controlled by both gravitational and capillary forces.

Capillarity results from two forces: the mutual attraction (cohesion) between water molecules and the molecular attraction (adhesion) between water and different solid materials. As a consequence of these forces, water will rise in small-diameter glass tubes to a height  $h_c$  above the water level in a large container (1).

Most pores in granular materials are of capillary size, and, as a result, water is pulled upward into a capillary fringe above the water table in the same manner that water would be pulled up into a column of sand whose lower end is immersed in water (2).

## APPROXIMATE HEIGHT OF CAPILLARY RISE ( $h_c$ ) IN GRANULAR MATERIALS

Material	Rise (mm)
Sand:	
Coarse -----	125
Medium -----	250
Fine -----	400
Silt -----	1,000

Steady-state flow of water in the unsaturated zone can be determined from a modified form of Darcy's law. Steady state in this context refers to a condition in which the moisture content remains constant, as it would, for example, beneath a waste-disposal pond whose bottom is separated from the water table by an unsaturated zone.

Steady-state unsaturated flow ( $Q$ ) is proportional to the effective hydraulic conductivity ( $K_e$ ), the cross-sectional area ( $A$ ) through which the flow occurs, and gradients due to both capillary forces and gravitational forces. Thus,

$$Q = K_e A \left( \frac{h_c - z}{z} \right) \pm \left( \frac{dh}{dl} \right) \quad (1)$$

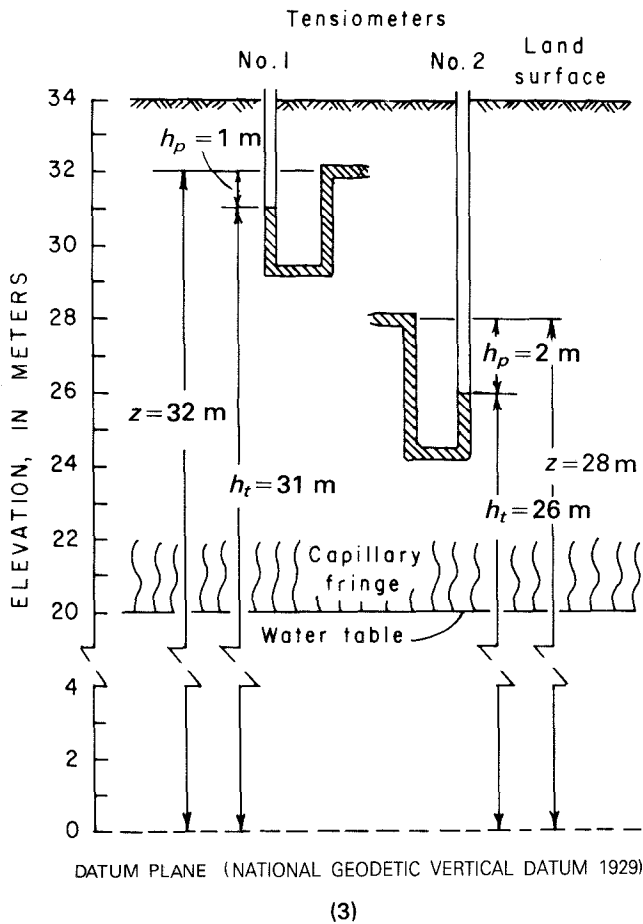
where  $Q$  is the quantity of water,  $K_e$  is the hydraulic conductivity under the degree of saturation existing in the unsaturated zone,  $(h_c - z)/z$  is the gradient due to capillary (surface tension) forces, and  $dh/dl$  is the gradient due to gravity.

The plus or minus sign is related to the direction of movement—plus for downward and minus for upward. For movement in a vertical direction, either up or down, the gradient due to gravity is 1/1, or 1. For lateral (horizontal) movement in the unsaturated zone, the term for the gravitational gradient can be eliminated.

The capillary gradient at any time depends on the length of the water column ( $z$ ) supported by capillarity in relation to the maximum possible height of capillary rise ( $h_c$ ) (2). For example, if the lower end of a sand column is suddenly submerged in water, the capillary gradient is at a maximum, and the rate of rise of water is fastest. As the wetting front advances up the column, the capillary gradient declines, and the rate of rise decreases (2).

The capillary gradient can be determined from tensiometer measurements of hydraulic pressures. To determine the gradient, it is necessary to measure the negative pressures ( $h_p$ ) at two levels in the unsaturated zone, as sketch 3 shows. The equation for total head ( $h_t$ ) is

$$h_t = z + h_p \quad (2)$$



where  $z$  is the elevation of a tensiometer. Substituting values in this equation for tensiometer no. 1, we obtain

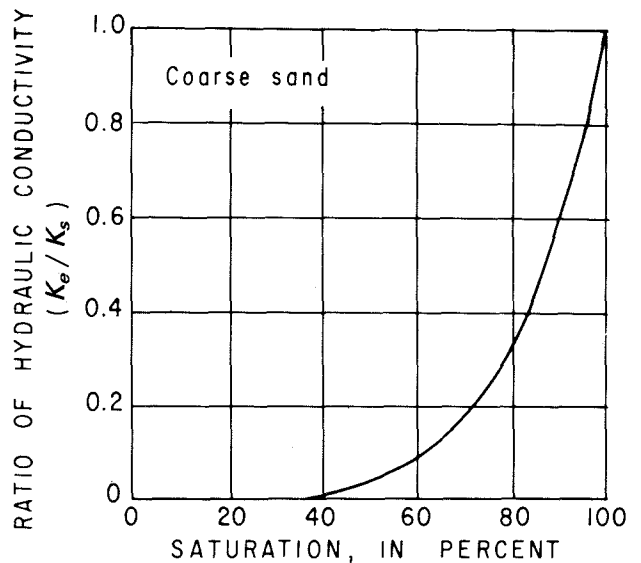
$$h_t = 32 + (-1) = 32 - 1 = 31 \text{ m}$$

The total head at tensiometer no. 2 is 26 m. The vertical distance between the tensiometers is 32 m minus 28 m, or 4 m. Because the combined gravitational and capillary hydraulic

gradient equals the head loss divided by the distance between tensiometers, the gradient is

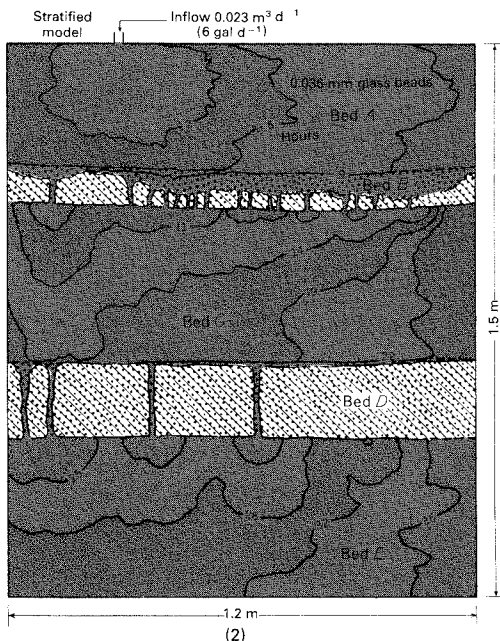
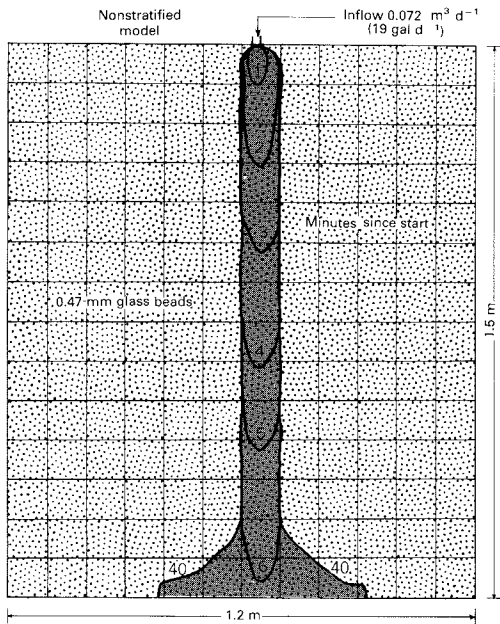
$$\frac{h_L}{L} = \frac{h_{t(1)} - h_{t(2)}}{z_{(1)} - z_{(2)}} = \frac{31 - 26}{32 - 28} = \frac{5 \text{ m}}{4 \text{ m}} = 1.25$$

This gradient includes both the gravitational gradient ( $dh/dl$ ) and the capillary gradient ( $[h_c - z]/z$ ). Because the head in tensiometer no. 1 exceeds that in tensiometer no. 2, we know that flow is vertically downward and that the gravitational gradient is 1/1, or 1. Therefore, the capillary gradient is 0.25  $\text{m m}^{-1}$  ( $1.25 - 1.00$ ).



The effective hydraulic conductivity ( $K_e$ ) is the hydraulic conductivity of material that is not completely saturated. It is thus less than the (saturated) hydraulic conductivity ( $K_s$ ) for the material. Sketch 4 shows the relation between degree of saturation and the ratio of saturated and unsaturated hydraulic conductivity for coarse sand. The hydraulic conductivity ( $K_s$ ) of coarse sand is about 60  $\text{m d}^{-1}$ .

# STRATIFICATION AND UNSATURATED FLOW



EXPLANATION

Areas remaining dry after 38 hours of inflow

Most sediments are deposited in layers (beds) that have a distinct grain size, sorting, or mineral composition. Where adjacent layers differ in one of these characteristics or more, the deposit is said to be *stratified*, and its layered structure is referred to as *stratification*.

The layers comprising a stratified deposit commonly differ from one another in both grain size and sorting and, consequently, differ from one another in hydraulic conductivity. These differences in hydraulic conductivity significantly affect both the percolation of water across the unsaturated zone and the movement of ground water.

In most areas, the unsaturated zone is composed of horizontal or nearly horizontal layers. The movement of water, on the other hand, is predominantly in a vertical direction. In many ground-water problems, and especially in those related to the release of pollutants at the land surface, the effect of stratification on movement of fluids across the unsaturated zone is of great importance.

The manner in which water moves across the unsaturated zone has been studied by using models containing glass beads. One model (1) contained beads of a single size representing a nonstratified deposit, and another (2) consisted of five layers, three of which were finer grained and more impermeable than the other two. The dimensions of the models were about  $1.5 \text{ m} \times 1.2 \text{ m} \times 76 \text{ mm}$ .

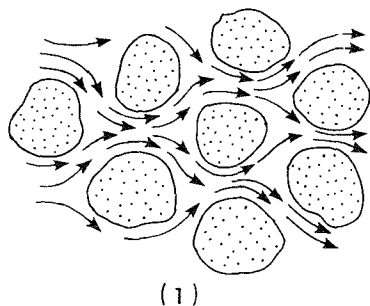
In the nonstratified model, water introduced at the top moved vertically downward through a zone of constant width to the bottom of the model (1). In the stratified model, beds A, C, and E consisted of silt-sized beads (diameters of 0.036 mm) having a capillary height ( $h_c$ ) of about 1,000 mm and a hydraulic conductivity ( $K$ ) of  $0.8 \text{ m d}^{-1}$ . Beds B and D consisted of medium-sand-sized beads (diameters of 0.47 mm) having a capillary height of about 250 mm and a hydraulic conductivity of  $82 \text{ m d}^{-1}$ .

Because of the strong capillary force and the low hydraulic conductivity in bed A, the water spread laterally at almost the same rate as it did vertically, and it did not begin to enter bed B until 9 hours after the start of the experiment. At that time, the capillary saturation in bed A had reached a level where the unsatisfied (remaining) capillary pull in bed A was the same as that in bed B. In other words,  $z$  in bed A at that time equaled  $1,000 \text{ mm} - 250 \text{ mm}$ , or  $750 \text{ mm}$ . (For a definition of  $z$ , see "Capillarity and Unsaturated Flow.")

Because the hydraulic conductivity of bed B was 100 times that of bed A, water moved across bed B through narrow vertical zones. We can guess that the glass beads in these zones were packed somewhat more tightly than those in other parts of the beds.

# SATURATED FLOW AND DISPERSION

Dispersion in a granular deposit



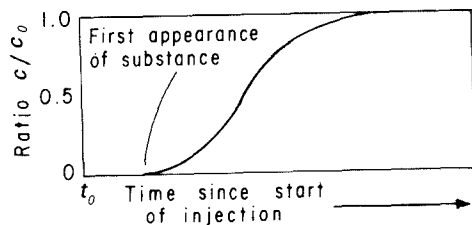
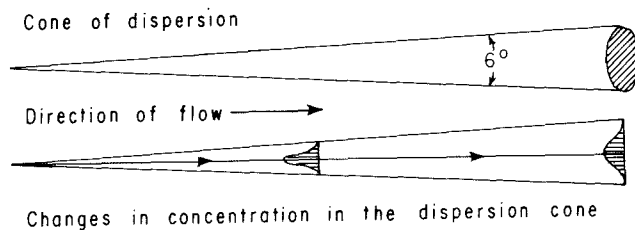
In the saturated zone, all interconnected openings are full of water, and the water moves through these openings in the direction controlled by the hydraulic gradient. Movement in the saturated zone may be either laminar or turbulent. In *laminar flow*, water particles move in an orderly manner along streamlines. In *turbulent flow*, water particles move in a disordered, highly irregular manner, which results in a complex mixing of the particles. Under natural hydraulic gradients, turbulent flow occurs only in large openings such as those in gravel, lava flows, and limestone caverns. Flows are laminar in most granular deposits and fractured rocks.

In laminar flow in a granular medium, the different streamlines converge in the narrow necks between particles and diverge in the larger interstices (1). Thus, there is some intermingling of streamlines, which results in transverse dispersion—that is, dispersion at right angles to the direction of ground-water flow. Also, differences in velocity result from friction between the water and the rock particles. The slowest rate of movement occurs adjacent to the particles, and the fastest rate occurs in the center of pores. The resulting dispersion is longitudinal—that is, in the direction of flow.

Danel (1953) found that dye injected at a point in a homogeneous and isotropic granular medium dispersed laterally in the shape of a cone about  $6^\circ$  wide (2). He also found that the concentration of dye over a plane at any given distance from the inlet point is a bell-shaped curve similar to the normal probability curve. Because of transverse and longitudinal dispersion, the peak concentration decreased in the direction of flow.

The effect of longitudinal dispersion can also be observed from the change in concentration of a substance (C) downstream from a point at which the substance is being injected constantly at a concentration of  $C_0$ . The concentration rises slowly at first as the "fastest" streamlines arrive and then rises rapidly until the concentration reaches about  $0.7 C_0$ , at which point the rate of increase in concentration begins to decrease (3).

Dispersion is important in the study of ground-water pollution. However, it is difficult to measure in the field because the rate and direction of movement of wastes are also affected by stratification, ion exchange, filtration, and other conditions and processes. Stratification and areal differences in lithology and other characteristics of aquifers and confining beds actually result in much greater lateral and longitudinal dispersion than that measured by Danel for a homogeneous and isotropic medium.



# GROUND-WATER MOVEMENT AND TOPOGRAPHY

It is desirable, wherever possible, to determine the position of the water table and the direction of ground-water movement. To do so, it is necessary to determine the altitude, or the height above a datum plane, of the water level in wells. However, in most areas, general but very valuable conclusions about the direction of ground-water movement can be derived from observations of land-surface topography.

Gravity is the dominant driving force in ground-water movement. Under natural conditions, ground water moves "down-hill" until, in the course of its movement, it reaches the land surface at a spring or through a seep along the side or bottom of a stream channel or an estuary.

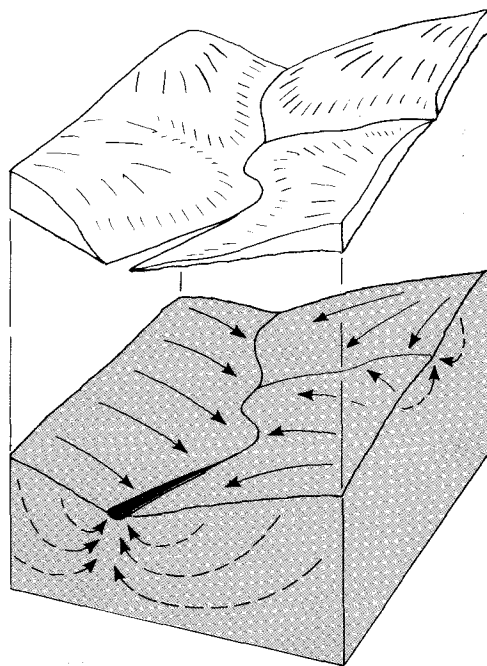
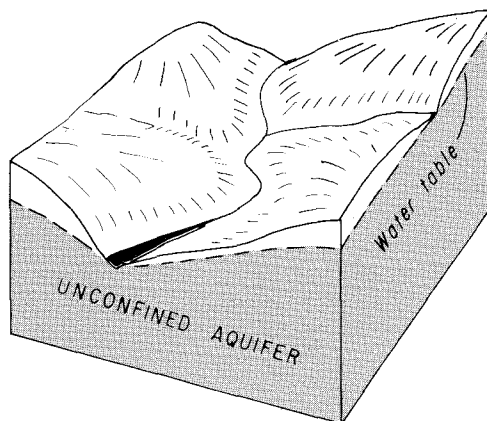
Thus, ground water in the shallowest part of the saturated zone moves from interstream areas toward streams or the coast. If we ignore minor surface irregularities, we find that the slope of the land surface is also toward streams or the coast. The depth to the water table is greater along the divide between streams than it is beneath the flood plain. In effect, the water table usually is a subdued replica of the land surface.

In areas where ground water is used for domestic and other needs requiring good-quality water, septic tanks, sanitary landfills, waste ponds, and other waste-disposal sites should not be located uphill from supply wells.

The potentiometric surface of confined aquifers, like the water table, also slopes from recharge areas to discharge areas. Shallow confined aquifers, which are relatively common along the Atlantic Coastal Plain, share both recharge and discharge areas with the surficial unconfined aquifers. This sharing may not be the case with the deeper confined aquifers. The principal recharge areas for these are probably in their outcrop areas near the western border of the Coastal Plain, and their discharge areas are probably near the heads of the estuaries along the major streams. Thus, movement of water through these aquifers is in a general west to east direction, where it has not been modified by withdrawals.

In the western part of the conterminous United States, and especially in the alluvial basins region, conditions are more variable than those described above. In this area, streams flowing from mountain ranges onto alluvial plains lose water to the alluvial deposits; thus, ground water in the upper part of the saturated zone flows down the valleys and at an angle away from the streams.

Ground water is normally hidden from view; as a consequence, many people have difficulty visualizing its occurrence and movement. This difficulty adversely affects their ability to understand and to deal effectively with ground-water-related problems. This problem can be partly solved



Arrows show direction of ground-water movement

through the use of flow nets, which are one of the most effective means yet devised for illustrating conditions in ground-water systems.



# GROUND-WATER FLOW NETS

Flow nets consist of two sets of lines. One set, referred to as *equipotential lines*, connects points of equal head and thus represents the height of the water table, or the potentiometric surface of a confined aquifer, above a datum plane. The second set, referred to as *flow lines*, depicts the idealized paths followed by particles of water as they move through the aquifer. Because ground water moves in the direction of the steepest hydraulic gradient, flow lines in isotropic aquifers are perpendicular to equipotential lines—that is, flow lines cross equipotential lines at right angles.

There are an infinite number of equipotential lines and flow lines in an aquifer. However, for purposes of flow-net analysis, only a few of each set need be drawn. Equipotential lines are drawn so that the drop in head is the same between adjacent pairs of lines. Flow lines are drawn so that the flow is equally divided between adjacent pairs of lines and so that, together with the equipotential lines, they form a series of “squares.”

Flow nets not only show the direction of ground-water movement but can also, if they are drawn with care, be used to estimate the quantity of water in transit through an aquifer. According to Darcy’s law, the flow through any “square” is

$$q = Kbw \left( \frac{dh}{dl} \right) \quad (1)$$

and the total flow through any set or group of “squares” is

$$Q = nq \quad (2)$$

where  $K$  is hydraulic conductivity,  $b$  is aquifer thickness at the midpoint between equipotential lines,  $w$  is the distance be-

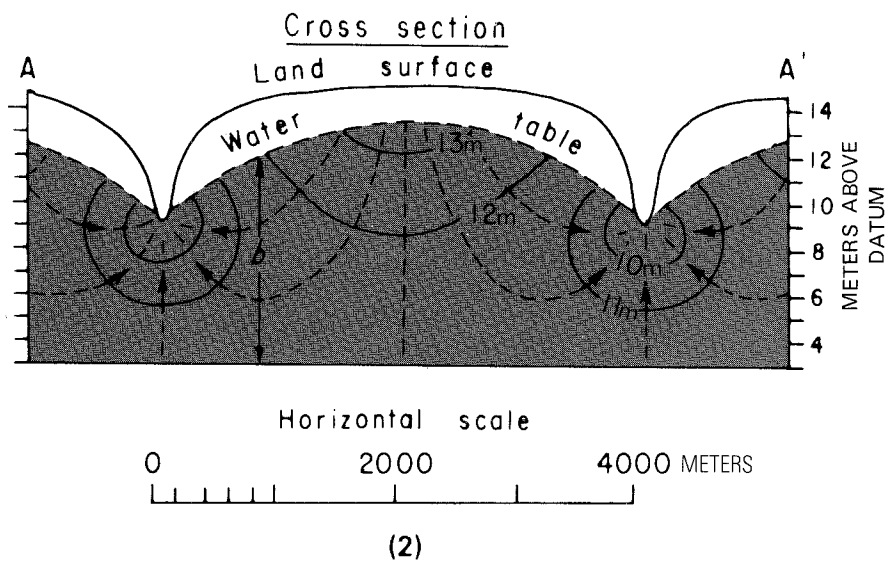
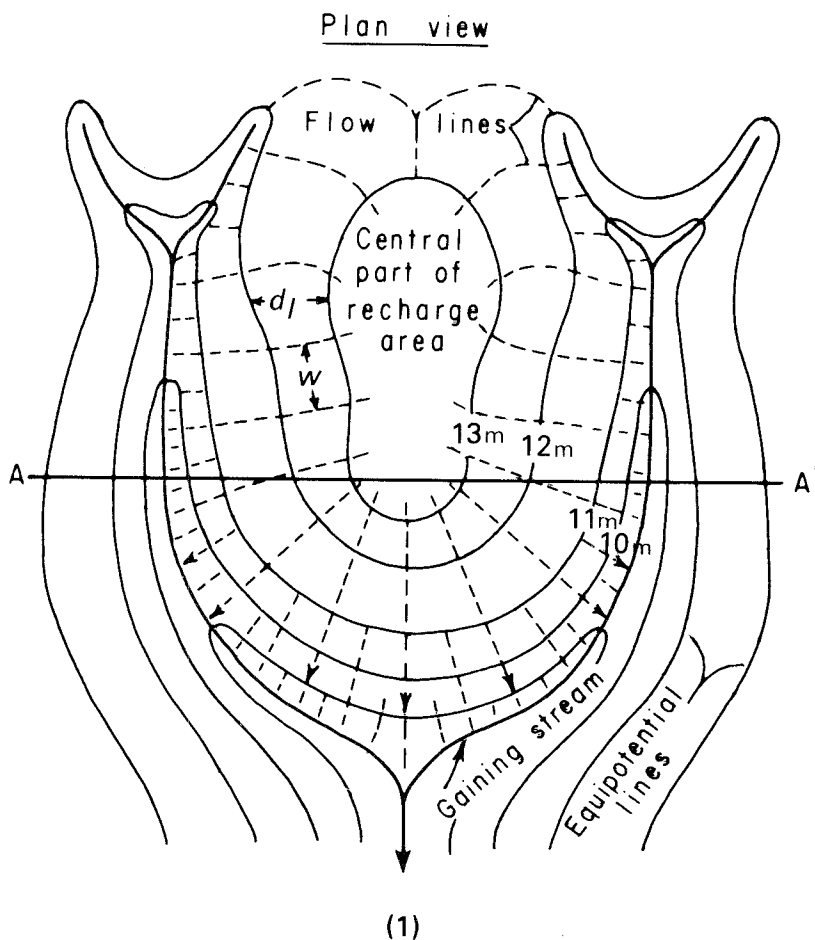
tween flow lines,  $dh$  is the difference in head between equipotential lines,  $dl$  is the distance between equipotential lines, and  $n$  is the number of squares through which the flow occurs.

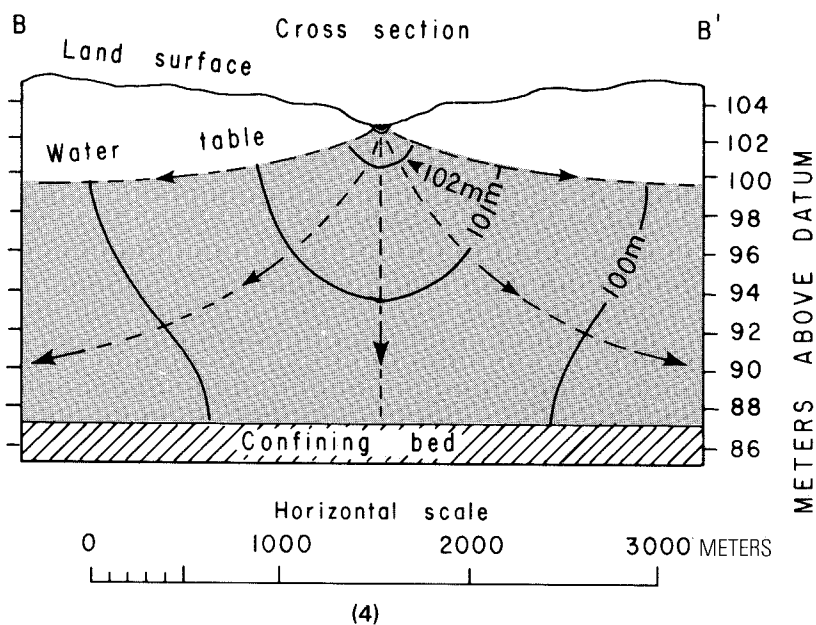
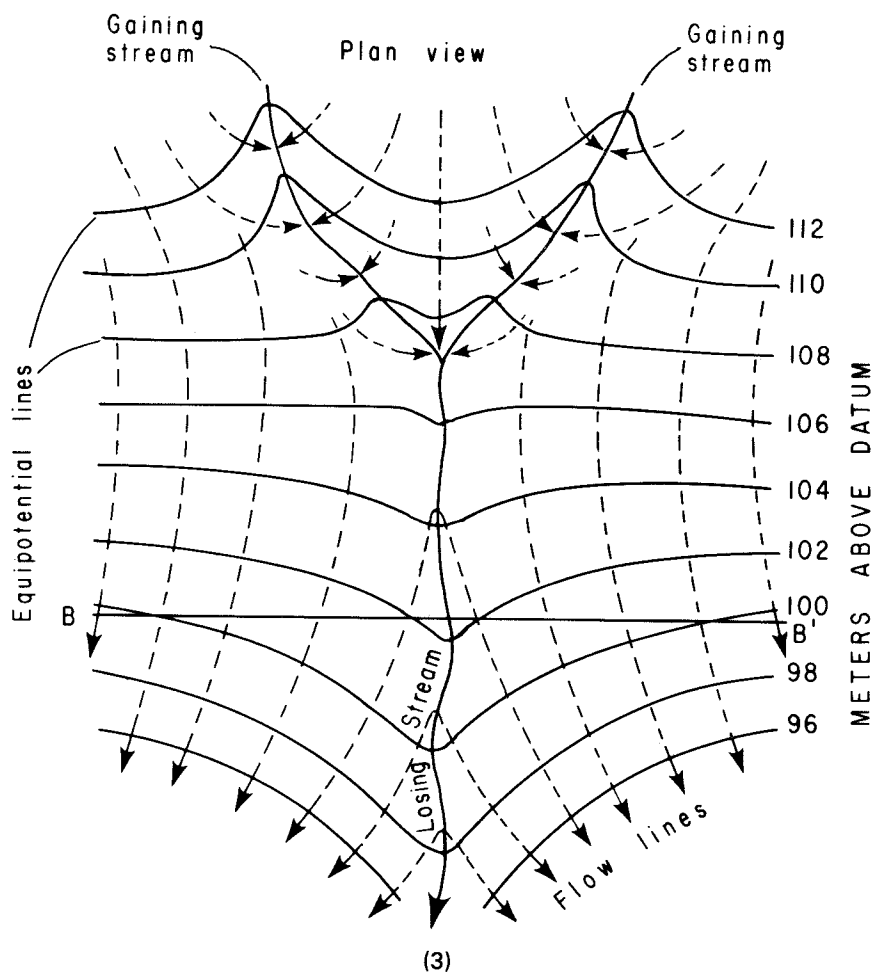
Drawings 1 and 2 show a flow net in both plan view and cross section for an area underlain by an unconfined aquifer composed of sand. The sand overlies a horizontal confining bed, the top of which occurs at an elevation 3 m above the datum plane. The fact that some flow lines originate in the area in which heads exceed 13 m indicates the presence of recharge to the aquifer in this area. The relative positions of the land surface and the water table in sketch 2 suggest that recharge occurs throughout the area, except along the stream valleys. This suggestion is confirmed by the fact that flow lines also originate in areas where heads are less than 13 m.

As sketches 1 and 2 show, flow lines originate in recharge areas and terminate in discharge areas. Closed contours (equipotential lines) indicate the central parts of recharge areas but do not normally indicate the limits of the areas.

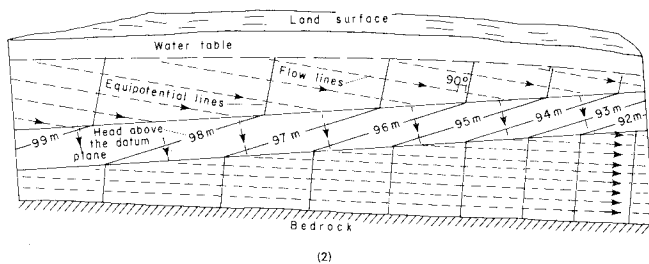
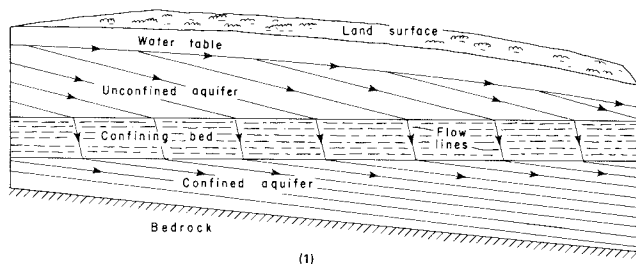
In the cross-sectional view in sketch 2, heads decrease downward in the recharge area and decrease upward in the discharge area. Consequently, the deeper a well is drilled in a recharge area, the lower the water level in the well stands below land surface. The reverse is true in discharge areas. Thus, in a discharge area, if a well is drilled deeply enough in an unconfined aquifer, the well may flow above land surface. Consequently, a flowing well does not necessarily indicate artesian conditions.

Drawings 3 and 4 show equipotential lines and flow lines in the vicinity of a stream that gains water in its headwaters and loses water as it flows downstream. In the gaining reaches, the equipotential lines form a V pointing upstream; in the losing reach, they form a V pointing downstream.





# GROUND-WATER MOVEMENT AND STRATIFICATION



Nearly all ground-water systems include both aquifers and confining beds. Thus, ground-water movement through these systems involves flow not only *through* the aquifers but also *across* the confining beds (1).

The hydraulic conductivities of aquifers are tens to thousands of times those of confining beds. Thus, aquifers offer the least resistance to flow, the result being that, for a given rate of flow, the head loss per unit of distance along a flow line is tens to thousands of times less in aquifers than it is in confining beds. Consequently, lateral flow in confining beds usually is negligible, and flow lines tend to "concentrate" in aquifers and be parallel to aquifer boundaries (2).

Differences in the hydraulic conductivities of aquifers and confining beds cause a refraction or bending of flow lines at their boundaries. As flow lines move from aquifers into confining beds, they are refracted toward the direction perpendicular to the boundary. In other words, they are refracted in the direction that produces the shortest flow path in the confining bed. As the flow lines emerge from the confining bed, they are refracted back toward the direction parallel to the boundary (1).

The angles of refraction (and the spacing of flow lines in adjacent aquifers and confining beds) are proportional to the differences in hydraulic conductivities ( $K$ ) (3) such that

$$\frac{\tan \theta_1}{\tan \theta_2} = \frac{K_1}{K_2}$$

In cross section, the water table is a flow line. It represents a bounding surface for the ground-water system; thus, in the development of many ground-water flow equations, it is assumed to be coincident with a flow line. However, during periods when recharge is arriving at the top of the capillary fringe, the water table is also the point of origin of flow lines (1).

The movement of water through ground-water systems is controlled by the vertical and horizontal hydraulic conductivities and thicknesses of the aquifers and confining beds and the hydraulic gradients. The maximum difference in head exists between the central parts of recharge areas and discharge areas. Because of the relatively large head loss that occurs as water moves across confining beds, the most vigorous circulation of ground water normally occurs through the shallowest aquifers. Movement becomes more and more lethargic as depth increases.

The most important exceptions to the general situation described in the preceding paragraph are those systems in which one or more of the deeper aquifers have transmissivities significantly larger than those of the surficial and other shallower aquifers. Thus, in eastern North Carolina, the Castle Hayne Limestone, which occurs at depths ranging from about 10 to about 75 m below land surface, is the dominant aquifer because of its very large transmissivity, although it is overlain in most of the area by one or more less permeable aquifers.

